

ALTITUDE DISTRIBUTION OF IONOSPHERIC NONHOMOGENEITIES
RESPONSIBLE FOR DIFFUSE REFLECTIONS OF SIGNALS
ABOVE DUSHANBE

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16. Abstract The question of the altitude distribution of small-scale ($1 < 1-5$ km) nonhomogeneities of the ionosphere at various latitudes is of considerable interest for understanding the nature of their origin. One of the few methods for studying this distribution at altitudes below the height z_0 of the F_2 layer maximum is the method based on analysis of the frequency dependence of the magnitude of the diffuse reflections of signals on vertical sounding ionograms (the F_{spread} phenomenon). [1,2].			
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(Presented by P. B. Babadzhanov, Academician of the
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The question of the altitude distribution of small-scale ($l < 1-5 \text{ km}$) nonhomogeneities of the ionosphere at various latitudes is of considerable interest for understanding the nature of their origin. One of the few methods for studying this distribution at altitudes below the height z_0 of the F_2 layer maximum is the method based on analysis of the frequency dependence of the magnitude of the diffuse reflections of signals on vertical sounding ionograms (the F_{spread} phenomenon) [1,2].

The diffraction theory suggested in [1] makes it possible to describe quantitatively certain types of F_{spread} phenomena and estimate the concentration fluctuations $\frac{\langle (\Delta N)^2 \rangle}{\langle N \rangle}$ in the nonhomogeneities at various altitudes.

According to this theory, certain types of F_{spread} , characterized by quasisymmetric spreading relative to z of the trace on the ionogram about to its mean value (and of the imperfect structure of the diffuse trace), are associated with the geometric-

* Numbers in margin indicate pagination in original foreign text.

optic fluctuations of the wave propagation time in the layer with statistical nonhomogeneities. In this case, the trace width

$$\Delta z \approx c \frac{8\sqrt{2} l_{\text{eff}} \sqrt{\langle S^2 \rangle}}{\omega \sqrt{\ln\left(\frac{8 l_{\text{eff}}}{l}\right)}},$$

$$\langle S^2 \rangle \approx \frac{\omega^2}{c^2} \sqrt{\pi} l_{\text{eff}} \ln\left(\frac{8 l_{\text{eff}}}{l}\right) T_\varepsilon, \quad (1)$$

where c is the speed of light, l is the nonhomogeneity dimension T_ε is a function depending on the altitude distribution of

$\frac{\Delta N}{N}$ and l , for example, (for $l(z) = \text{const}$ and $\frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}(z) = \text{const}$)

$$T_\varepsilon = \frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}, \text{ while for } l(z) = \text{const} \text{ and } \frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}(z) = \text{const}, T_\varepsilon = \frac{\omega_{\text{cr}} \langle (\Delta N)^2 \rangle}{\omega \langle N^2 \rangle} l_{\text{eff}}$$

is the effective layer thickness, which for the parabolic layer model with thickness Z is

$$l_{\text{eff}} = \frac{\omega^2 Z_m}{2 \omega_{\text{cr}}^2 \sqrt{1 - \left(\frac{\omega}{\omega_{\text{cr}}}\right)^2}} \quad (2)$$

($\omega = 2\pi f$, ω_{cr} is the layer critical frequency).

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Calculations of (1) for two models of the nonhomogeneities present in the parabolic layer show that $\Delta z \sim \omega^{-3}$ for $\frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}(z) = \text{const}$ and $\Delta z \sim \omega^{-1}$ for $\frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}(z) = \text{const}$. The second model, in contrast with the first, is characterized by sharp decrease of the relative concentration fluctuations with height $\left(\frac{\Delta N}{N} \sim \frac{1}{N}\right)$. Naturally, for large decrease of $\frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}$ with height, the quantity $\Delta z \sim \omega^{-n}$, where $n < 1$, while for increase of $\frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}$ with height, the frequency dependence exponent n will be larger than 3.

Regardless of the applicability of the results of [1] to the real atmosphere, we can expect that slower decrease of $\frac{\langle (\Delta N)^2 \rangle}{\langle N^2 \rangle}$ with height should be characterized, for unchanged ionospheric layer form, by greater dependence of Δz on frequency (although the question of correspondence of a given n with the quantitative nonhomogeneity altitude model remains open).

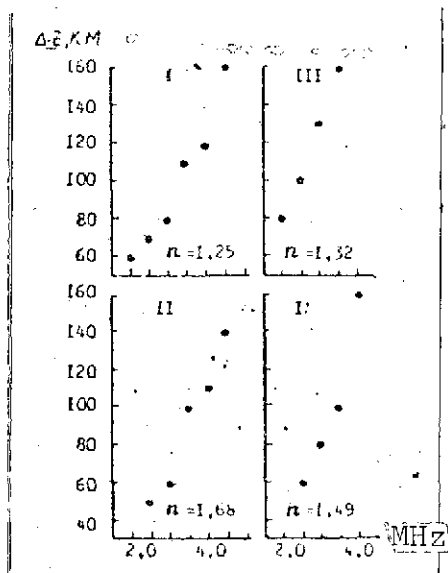


Figure 1. Relation $\Delta z \sim f^{(n)}$ from Dushanbe station data for I - IV 1971

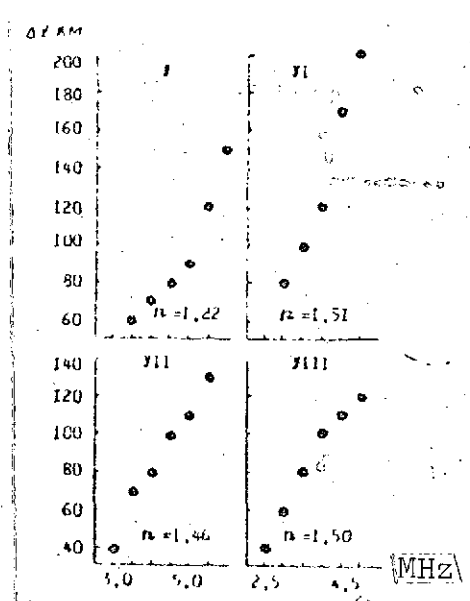


Figure 2. Relation $\Delta z \sim f^{(n)}$ from Dushanbe station data for V - VIII 1971

In this connection it is interesting to examine the experimental data on Δz frequency dependence obtained at the middle-latitude station of the Institute of Astrophysics of the Academy of Sciences Tadzhik SSR (at Dushanbe). The results presented are from observations made in 1971.

For better correspondence with the results of [1] we analyzed ionograms with diffusivity symmetric relative to the primary trace. Moreover, in order to exclude apparatus errors, we selected for analysis ionograms similar to the calibration recordings. In the study we used more than 700 ionograms for determining Δz . The diffusivity origins on the ionograms are taken as the origin for the Δz measurements. The last Δz value from the ionogram is determined from the condition $f < f_0$, since the diffusivity thickness dependence on frequency is disrupted near the critical frequency; f is the frequency at which Δz is measured and f_0 is the F_2 layer critical frequency. The analysis /27

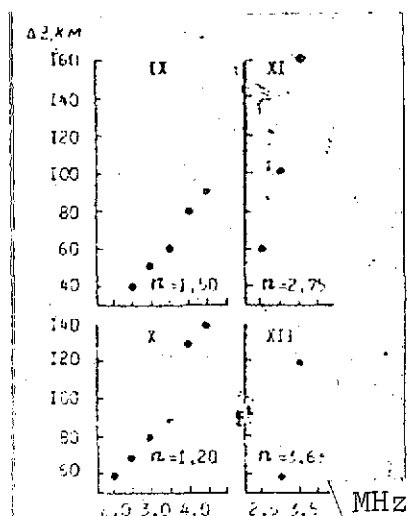


Figure 3. Relation $\Delta z \sim f^n$ for Dushanbe station data for IX-XII 1971.

results are shown in Figures 1, 2 and 3. The values of the effective exponent of the frequency dependence $\Delta z \sim f^n$ are indicated in the figures. The value of Δz in the summer and solstice months at fixed frequencies is larger than in the winter months. In all cases the quantity $n < 3$. This fact, which agrees with the results of observations at another middle-latitude station (Ashkhabad), indicates that at middle latitudes the quantity

$\frac{\langle (\Delta n)^2 \rangle}{\langle n^2 \rangle}$ at altitudes $z < z_0$ decreases with altitude increase (we are considering nonhomogeneities with dimensions $1 \sim 1-5 \text{ km}$)

Thus the source of small-scale nonhomogeneity formation must be more effective (at the indicated latitudes) at low ionospheric F layer altitudes. At the same time, since for all the observation months n was greater than unity, we can apparently state that in the ionosphere there is not simple transport of isolated nonhomogeneities from lower to higher altitudes (in this case $\langle (\Delta n)^2 \rangle$ should remain constant).

The variations of the quantity n from month to month are interesting. Except for two months (November and December, 1971 — 2.75 and 3.63) the exponent n varied comparatively little — from 1.2 to 1.7. At the present time it is difficult to judge what such variations could be associated with, since from the averaged data no dependence of n on solar (Wolf number) or magnetic (monthly-mean k index) activity could be detected. However, the latter may be associated with the particular selection

of the F_{spread} observation data. It seems to us that in the future it is advisable to classify the ionograms on the basis of the degree of Δz dependence on the frequency and establish the connection between this classification and the commonly used indices of F_{spread} intensity [2,3]. The question of variation of n as a function of the ionospheric and magnetic disturbance level is of great interest.

In conclusion we shall estimate the quantities $\frac{\langle (\Delta n)^2 \rangle}{\langle n^2 \rangle}$ for some characteristic $|\Delta z(\omega)|$ curves (Figures 1, 2, and 3). Since for November, 1971 $n \approx 3$, we can use the model $\frac{\langle (\Delta n)^2 \rangle}{\langle n^2 \rangle}(z) = \text{const}$, while for January, 1971, when $n \approx 1$, we can use the model $\langle (\Delta n)^2 \rangle(z) = \text{const}$.

Then, using (1), (2), we have

$$\Delta Z \approx \frac{8\sqrt{2} L_{\text{eff}}^{3/2}}{\sqrt{f}} T_1^{1/2}$$

Taking for definiteness $l = 2$ km, for November, 1971, we have $\frac{\Delta n}{n} = \sqrt{\frac{\langle (\Delta n)^2 \rangle}{\langle n^2 \rangle}} = T_1^{1/2}$, which is equal to $2.7 \cdot 10^{-2}$, $4.3 \cdot 10^{-2}$ and $5.6 \cdot 10^{-2}$ respectively, at altitudes of 350, 300, and 250 km.

For January, 1971, $T_1^{1/2} = \frac{cn}{\omega} \sqrt{\frac{\langle (\Delta n)^2 \rangle}{\langle n^2 \rangle}}$, then we obtain a value of $\frac{\Delta n}{n}$ at altitudes 300-350 km of the same order as for

November, 1971. We note that the quantities $\frac{\Delta n}{n}$ depend comparatively weakly on nonhomogeneity size. The quantity $\frac{\Delta n}{n}$ is considerably more sensitive to the parameter L_{eff} , which determines the form of the mean layer.

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